OC 3570

OPERATIONAL OCEANOGRAPHY AND METEOROLOGY

PROJECT REPORT

MEASUREMENT OF THE CALIFORNIA LOW-LEVEL COASTAL JET USING RAWINDSONDS

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1. INTRODUCTION

A coastal jet (CJ) is a low-tropospheric wind feature driven by the pressure gradient produced by a sharp contrast between high temperatures over land and lower temperatures over sea (Cross 2003).

The CJ have been identified and studied in several areas of the world, where such a land-sea temperature contrast exist; off the coast of Somalia; near Lima, Peru; off the Mediterranean coast of Spain; in the Southwest coast of Africa, or on the China Sea coast. Nevertheless, the California CJ is probably the most studied CJ in the world, with a relative abundance of papers in the literature specifically about the California CJ features, or based on it([1]).

Low-level CJs have a notorious impact on coastal areas. Climatologically they are associated with upwelling processes, which create pools of cold water near the coast. The major coastal fishing grounds in the world are usually in areas of upwelling, and the abundance of fish at the surface is supported by the upwelling of nutrient-rich waters from deeper levels. The effect of this upwelled water to the fishing industry and to the habitat of an enormous diversity of marine life is of paramount importance, and has led to numerous studies in this field.

The climate of the along-coast areas, where wind-driven upwelling takes place, tends to be mild, with the cold ocean waters having a great influence in the air temperature on the shore near the coast.

[1] Most of the knowledge about the California CJ derives from the data collected during the two phases of the Coastal Ocean Dynamics Experiment (CODE) that took place during the 1981 and 1982 upwelling seasons (Beardsley et al. 1987).

Littoral areas are usually densely populated, and often airports are built in areas where a CJ may occur. Thus, aviation operations are deeply influenced by this weather feature, which has a great impact on the take off and landing of airplanes.

Modern naval operations are more and more often conducted in littoral waters (Forward from the Sea). When the high winds of the jet reach down to the surface, the potential exists for winds that could be dangerous to a variety of naval operations, mainly with aircrafts.

The purpose of this study is to check the viability of measuring a CJ features using rawinsondes as the only observational system, launched at sea from a Research Vessel (RV).

This report will focus initially on theoretical aspects of the California CJ, its forcing regime, and interactions with other mesoscale coastal processes that modulate its spatial and temporal regime. Further on it will describe the data measurements of the California CJ done on board the RV Point Sur of the coast of the Big Sur, from August 4th to August 7th, 2004.

The synoptic picture during the cruise was obtained using the 12-h Navy Operational Global Atmospheric Prediction System NOGAPS analysis generated at the U.S. Navy FNMOC (Fleet Numerical Meteorology and Oceanography Center). The Pennsylvania State University and Nacional Center for Atmospheric Research (PSU/NCAR) mesoscale model (known as MM5) was further used for verification and validation of the observations.

2. CALIFORNIA COASTAL JET STRUCTURE

The United States (US) West Coast, during the spring and summer months, is under the influence of the Pacific high, which is centered approximately 1000 Km west of the California coast, near 40° N (Beardsley et al. 1987), and a thermal low over the desert southwest (Fig. 1), which is generated by the increased solar insolation in the spring and summer. This climatological distribution of sea-level pressure gives rise to persistent northerly to northwesterly geostrophic coast-parallel winds over the eastern Pacific Ocean, along the California coast. Due to the presence of a strong low-level baroclinic structure, the pressure gradient is maximum at the coast and decreases both landward and seaward of the coastal boundary.

The persistent northwesterly flow along the sea surface forces coastal upwelling, which produces a very low sea surface temperature (SST) near the coast of California. This low SST cools the marine air near the surface. On the other hand, when subsiding air on the eastern flank of the eastern North Pacific High warms adiabatically and comes in contact with the cold air at the surface, a strong inversion is formed at the top of the marine boundary layer (MBL).

Due to the combination of a dramatic increase in SST and weakening of the synoptic-scale subsidence to the west, the inversion downward tilts toward the coast, and drops from an initial maximum elevation off shore of about 1500 m to about 300 m (Beardsley et al. 1987) (Fig. 2).

Through thermal wind considerations, a low-level wind maximum is generated, and the strongest cross-coast pressure gradient is at the surface. This yields the jet structure seen in Fig. 3 (Burk and Thompson 1996), which is the California CJ. The wind maxima is observed above the surface due to the deceleration of the wind closer to the ground that caused by friction.

The California CJ is a broad feature that can extend up to 500 Km offshore (Beardsley et al. 1987). Nevertheless, the jet core can be considerably compact, with a 20-40 km width in some places.

3. MESOSCALE INTERACIONS AND CJ VARIATIONS

3.1 Cross-coast and diurnal variations

The California CJ interactions with mesoscale coastal processes depend on aspects on the coastal topography not included in the broad baroclinic structure. Several studies on the California CJ showed that in some areas along the coast the inversion height has a diurnal variability related with a “daytime along-shore acceleration” (Beardsley et. al 1987) (Fig. 4).

In the morning, the sun starts heating the land surface, and a weak thermally-driven cross-coast circulation develops under the inversion. In the afternoon, right over the beach, the winds are parallel to the shore. Farther inland, on the narrow coastal plain (before the coastal mountains), the wind turns towards a more cross coast direction. A weak return flow develops above the inversion, which helps to depress the nearshore inversion even more.

As a result of this depression during the day, the inversion tilts downward to a minimum elevation that can be as low as 30 m (Winant et al. 1988) just over the beach. Inland from the beach, the inversion lifts very slowly and becomes diffuse.

The wind speed peaks right bellow the inversion, just off the beach. As the sun sets, the stability increases over land, due to radiative coolong, and the weak cross flow ceases. The land becomes again isolated from the high momentum along shore flow. The inversion lifts, and the surface winds diminish.

At night without the warming over land, the weak cross-coast flow ceases, the inversion lifts up to several hundred meters, and the surface winds die. The wind maximum in the nearshore region will be on the ocean side in the midmorning and on the land side in the midafternoon.

As Beardsley et al. mention, a real sea breeze is not present, since the flow is never really oriented in the cross coast (perpendicular direction). The disturbance is due to the intrusion of the high speed wind over the narrow coastal beach area.

3.2 Along-shore variations and topographic adjustment

An important aspect of the Califonia CJ is its along-coast variability, which impacts the cloud distribution, wind, and corresponding forcing on the upwelling of coastal waters. Observations during the CODE showed a well-defined jet maximum downwind of Point Arena and Point Sur (Fig. 5).

It as been hypothesized that this wind streak that occurred downstream of coastal obstructions, coupled with a decrease in the inversion height, was due to a two-layer supercritical hydraulic flow. Winant et al. presented a rather simplistic model of this supercritical channel flow to resolve the along-shore variation of the California CJ.

To explain how wind maxima occur downstream of points and capes, Winant et al. found it useful to consider the detailed structure of the flow in these regions based on the CORE observations. It was considered that atmospheric conditions during the spring and summer usually fall into one of three categories: (1) weak surface wind, (2) strong uniform flow, or (3) complex flow with large changes in wind speed and direction (Fig. 6 - Winant 1988).

Pattern 1 is characterized by a deep well-mixed MBL up to about 600 m and weak flow, with no particular along-shore variation. Patterns 2 and 3 are characterized by relatively strong northwesterly flow with different mesoscale coastal responses. Pattern 2 is characterized by a shallow well-mixed MBL with weak stratification above, which is typical of a post-frontal atmosphere. On the other hand, Pattern 3 is characterized by a well-mixed MBL, with a strong inversion above. The pressure distribution is most instructive and perfectly shows that the wind maximum in pattern 2 occurs in the region of largest pressure gradient, while the pattern 3 maximum and minimum occur in the regions of lowest and highest pressures respectively.

Winant et al. did not explain the mechanism that generates the mesoscale down-coast pressure gradient in pattern 2, but it is clear that this is due to thermal differences above the mixed layer (Nuss 2004([2])).

The strong low-level inversion in pattern 3 is usually below the coastal mountain top. In this two-layer structure, even small changes in the inversion base height will generate pressure changes at the surface. Assuming that the MBL height change is responsible for the pressure changes over the region, and then the low pressure regions are associated with low inversion base height and the high pressure with high inversion base height. Given this structure, high speed winds are directly associated with low inversion base height and low speed winds with high inversion base height. Winant et al. make the analogy of hydraulic flow through a pipe, with speed increasing when the diameter of the pipe decreases. This analogy considers the flow is trapped in a channel with mountains on one side, sea surface below, strong inversion above, and synoptic-scale pressure gradient on the other side and it is not far from reality.

In order to have the flow along the coast under such conditions, it is necessary to consider the behavior of the flow as governed by a Froude number defined as $Fr = \frac{V}{\sqrt{g' h}}$, where V is the flow speed in the layer and $\sqrt{g' h}$ is the phase speed of gravity waves from the shallow water equations, and $g' = g\frac{\Delta\rho}{\rho}$ is the reduced gravity. The phase speed also represents the potential energy needed to change the height of the interface. In this format, the Froude number,

[2] In-class notes.

represents the ratio of the speed of the flow in the layer to the phase speed of gravity waves propagating in the layer (Samelson 1992).

The abrupt change in the coastal orientation due to a cape or a point will, perturb the flow and generate gravity waves propagating in both upstream and downstream directions, or just downstream, depending upon whether the flow is subcritical or supercritical.

If the Froude number is less than one, the phase speed of gravity waves will be greater than the flow speed in the layer, and the waves can propagate both upstream and downstream (subcritical flow). Circulation changes are felt both upstream and downstream of the geometry change, due to gravity waves propagating radially away from the flow obstruction. When the Froude number is greater than one (supercritical flow), the phase speed of gravity waves is now less than the flow speed, and waves will travel downstream only. Since waves cannot propagate upstream, the obstruction affects only the downstream flow and adjusts it to the new boundary by depressing the inversion height, with a hydraulic jump farther downstream.

This hydraulic theory dictates that the height of the flow decreases when the velocity increases. If we neglect the Coriolis effect and friction in the momentum equation for the shallow water flow, we will get $\frac{dV}{dt} = g' \frac{\partial h}{\partial y}$, which defines the acceleration in terms of the pressure gradient in the y-direction, which is given by changes in inversion height for the shallow water equations. The momentum equation can also be integrated along a streamline to determine the energy balance given by a Bernoulli equation, which implies that the total energy (kinetic plus potential) must be conserved following a parcel $\frac{V^2}{2} + g^{'}h = cte$, relating the height changes of the inversion base to the flow speed changes following the trajectory. In order to have strong winds we will need a low inversion, and with a high inversion we will get weak winds.

As Winant et al. pointed out, the impact of supercritical adjustment on the MBL as it flows along a convex bend in the coastline is a spreading of the flow in (called an expansion fan) to follow the contours of the coastline. The result is a decrease of the inversion height and a corresponding increase in the wind speed from (3). As the coastal boundary extends into the flow, subsequently, a compression dictates convergence of the flow characteristic lines and necessitates a sudden jump in the inversion height with a corresponding decrease in the wind speed (Fig. 7). If we have a concave coastline orientation, this implies that the wave fronts should now occur upstream of their original position, which cannot occur physically. This results in a piling of wave fronts that produces an abrupt increase in the inversion height, which becomes a hydraulic jump. The flow speed decreases([3]).

[3] Samelson (1992) later pointed out that the Winant et al. model failed to predict the near-shore velocity maximum obtained in the CODE data, over-predicted the acceleration downstream of the point or cape, and didn't account properly for an observed deceleration upstream from the hydraulic jump. All of these shortcomings may be because friction, rotation, and upper-level pressure variations were neglected. The net impact of the inclusion of these additional effects in the Samelson model results in a better fit with the data from the CODE measurements.

3. OBSERVATIONAL ANALYSIS

The data used in this study was collected using rawinsonde measurements done on board the Research Vessel (RV) Point Sur during the 2004 OC3570 summer cruise (August 4^{th} to August 11^{th}). During this period 24 rawinsondes (Fig. 8) were launched. The ship followed a track along the California Cooperative Oceanic Fisheries Investigations (CALCOFI) lines 67, 70, 77 and 85. Lines 70 and 77 are parallel and perpendicular, respectively, to the Big Sur coast, where the California CJ is known to have a jet wind maximum, southward of Point Sur, due to topographic mesoscale interactions, resulting in a supercritical adjustment that leads to wind speed increase. This fact lead to the choice of using only the rawindsonde observations conducted along the CALCOFI lines 70 and 77, which narrowed the sondings to a number of 15 (numbers 4 to 18).

Since the cruise had multiple academic purposes, several oceanographic CTD measurements were being done in different stations along the mentioned CALCOFI lines. This situation considerably extended the observation period. In order to reduce the time scale of the study, it was decided not to use sondings 14 to 18, further reducing the number of observations to 10; sondings 4 to 8 in the coast parallel section (from here on designated as leg 1) and sondings 8 to 13 in the cross-coast section (from here on designated as leg 2). By reducing the number of observations used in the study, we were able to also reduce the time scale from a little more than 6 days to less than 59 hours (order of 1 day for each one of the legs)([4]).

As seen on Fig. 9, the synoptic situation remained relatively stable during the narrowed observational period, making the assumption of 1 plus 1 days for time scale perfectly valid. The pacific high pressure system was positioned approximately 1000 Km west of the California coast, near 40° N. The desert southwest thermal low was less intense, and its position was more northern than usual, which favored a coast parallel surface geostrophic wind offshore of the Big Sur coast, as depicted on Fig. 10. The conditions for having a coast parallel low-level CJ were apparently all present.

The length scale of each one of the observational legs, even if not ideal, was considered reasonable, based in previous MBL studies (Fig. 1).

Each one of the 10 sondings collected data in both upward and a downward measuring branches, and the measurements were done up to an average altitude of 3500 m. It was decided to use only the upward branch of the measurements and only until an altitude of 1500 m, since the capping inversion and the CJ are usually way under that height([5]).

The sondings plots for legs 1 and 2 are depicted in Fig. 11 and Fig. 12 (temperature and dew point temperature) respectively (sonding 8 is in showed in both figures), along with a graph where all temperature profiles are put together, offsetting them by 10°C, allowing a combined view of the profiles in each leg. In

[4] The duration of the observational period can be a problem, since the synoptic condition can vary, if the time scale is too large. As synoptic period (less than 6-h) might be the appropriate time scale to have in mind. Even so, the diurnal variations mentioned by Beardsley et al. cannot be sensed using this type of observational system and track in the sea.
[5] MATLAB was used to handle the data.

Fig. 13 and Fig. 14 the wind speed profile and wind direction (using a vertical "feather" graph, with the size of the wind barbs proportional to the wind speed) is depicted for each leg, respectively.

In the first leg all temperature profiles revealed a very weak or nearly inexistent inversion. The inversion height varied from sonding to sonding. On Fig. 13 we can se that there was a consistent wind shear near the surface and, with the exception of the sonding 4, the wind speed maximum was relatively coherent in terms of speed (14-18 m/s) and height (~500m). The variation of the wind direction with height was also consistent, showing minimum variation, with the exception, again, of sonding 4.

From this results we can say, regarding the first leg, that there was a strong evidence that the ship's track was running parallel to a low level coastal jet, inside a not very wide MBL.

As we moved towards the coast, and started the observations on leg 2, the inversion started to built and getting shallow, as seen in Fig. 12. After sonding 11 the inversion height rose about 50 m. Sonding 13 was not accountable, since the measurements were done already inside Port San Luis, with significant sheltering from the coastal topography. The wind speed maximum coincided with the inversion height in all sondings, showing evidence of flow trapping up above. The wind speed maximum showed a higher variability compared to the coast parallel leg, ranging from a maximum of 23 m/s on sonding 10 and a minimum of 12 m/s on sonding 11. This was somehow unexpected if we think only in terms of a direct relation between the decrease of the inversion height and the wind speed explained by the Bernoulli effect. But several other aspects, resulting from the interaction of the flow with Point Sur and the subsequent propagation of the gravity waves downstream, combined with local effects that can lead to the lowering of the inversion height (such has a pool of lower SST), or even a diurnal variation of the CJ, might have led to this situation.

From the measurements on leg 2, we can therefore conclude that a CJ was crossed, with a strong evidence of the jet core being in the vicinity of sonding 10. The squashing of the inversion height is correlated with the increase of the wind speed, if we account for marginal effects as mentioned above.

From the analysis of the data from legs 1 and 2 and the NOGAPS 12-h analysis, we can state that a low-level CJ feature was present along the Big Sur coast, and also that this jet interacted with the coastal topography in Point Sur, generating gravity waves and a consequent increase of the wind speed, due to supercritical adjustment.

4. MODEL VERIFICATION

To verify and validate the observations we used MM5 12-h forecast for 070000UTC, coinciding with 071900PDT. In Fig. 15 (a) (1010 mb and 24 m winds) we can see that the model forecasted a strong wind field parallel to the Big Sur coast, which was expected, given the NOGAPS synoptic picture and the cross-sectional observations. In Fig. 15 (b) (1000 mb isotachs) we could clearly see that the conclusion based on the observational analysis was correct. The model forecasted a flow interaction with the Point Sur coastal topography and a consequent jet streak downstream. The 1000 mb level roughly coincided with 200 m, which is consistent with the inversion height and the wind speed maximum in sondings 10 to 12.

In Fig. 16 a MM5 cross-coast crossection is depicted, Once again, the model forecasting matches the observations([6]). The isentropes are tilted towards the coast and show an inversion closer to what was showed in Fig. 12(d). In Fig. 16 the model forecasted a jet core vertical development ranging roughly from 1000 mb to 950 mb (200-300m), which was closer to the observations.

The model also forecasted, as can be seen in Fig. 17, that the only place in the all California coast where a CJ flow speed enhancement due to interaction with the topography could take place was in the Big Sur coastal area. This situation also verifies what was stated above, based on the NOGAPS synoptic picture.

5. CONCLUSIONS AN RECOMENDATIONS

It was proved possible the measurement of basic features of a low-level CJ with rawinsonds. It seamed feasible, with the proper planning, and with a higher number of observations, combined with a mesoscale model, that a thorough scientific study of a coastal jet can be done using this method. (But this was not the scope of this study.)

[6] In all the wind forecasts the model under predicts the observations.

The CJs are an important atmospheric feature in the littoral areas that can have an impact in the war fighter planning of air operations. Therefore, this type of study is to be encouraged during the summer cruises. There are low-level coastal jets in numerous areas of the globe, and without other means of observation, using a couple of rawinsonde observations, the METOC officer can assess the conditions and provide a better environmental support.

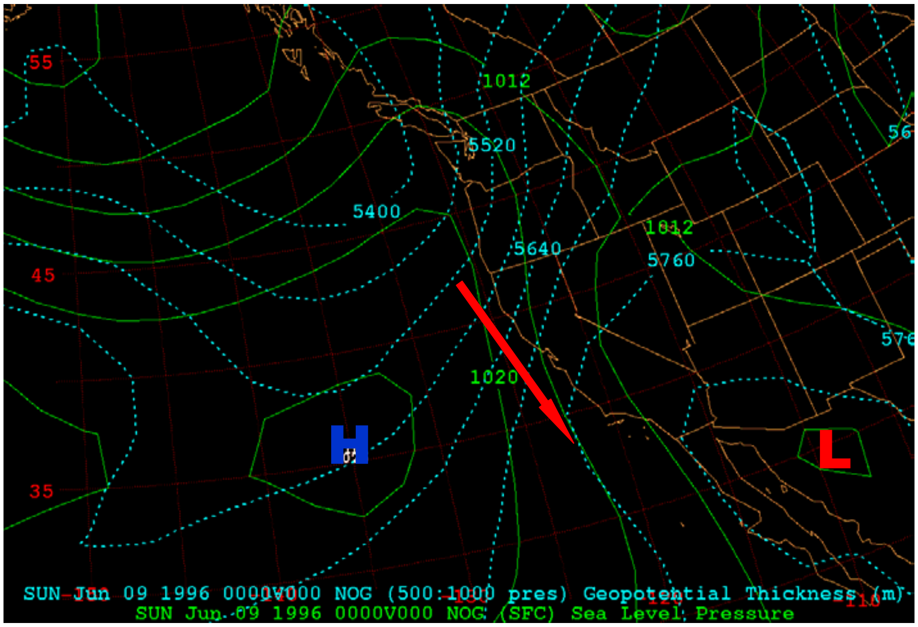


Fig.1. Example of surface air pressure and 500-100 mb geopotential thikness over the eastern north Pacific. (NOGAPS).

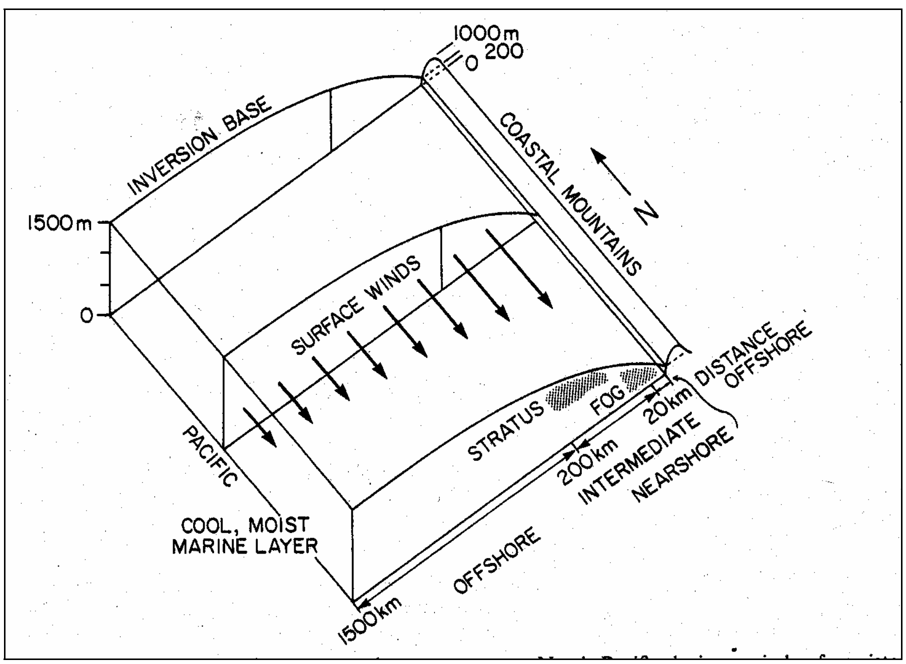


Fig. 2. Conceptual model of average lower atmosphere over eastern north Pacific. (From Beardsley et al. 1987).

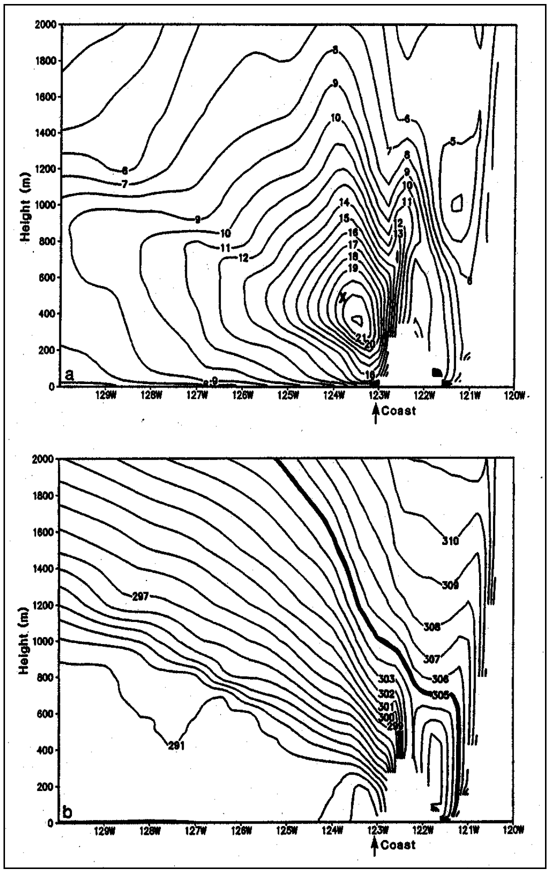


Fig. 3. Isotachs (m/s) and potential temperature cross-sections on the California Coast (From Burk and Thompson 1996).

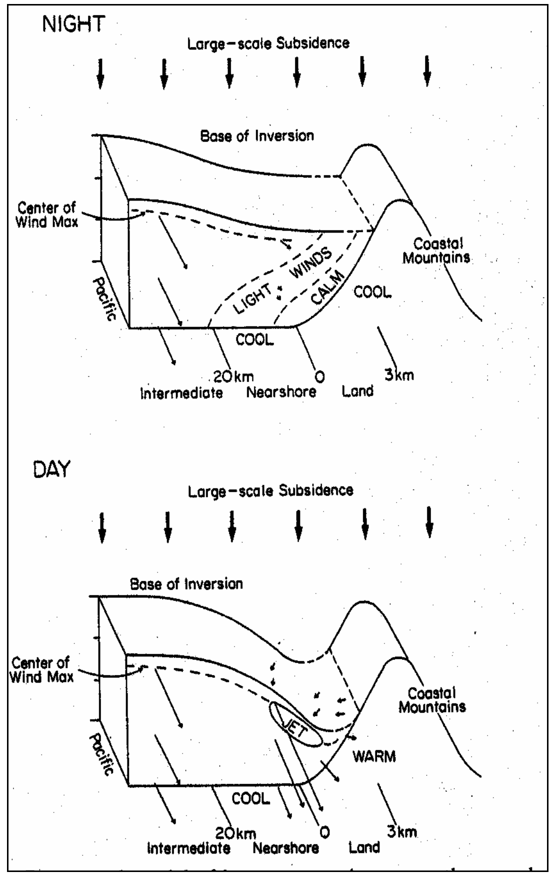


Fig. 4. Conceptual model of the lower atmosphere over the nearshore zone during the night (top) and day (bottom). (From Beardsley et. al 1987).

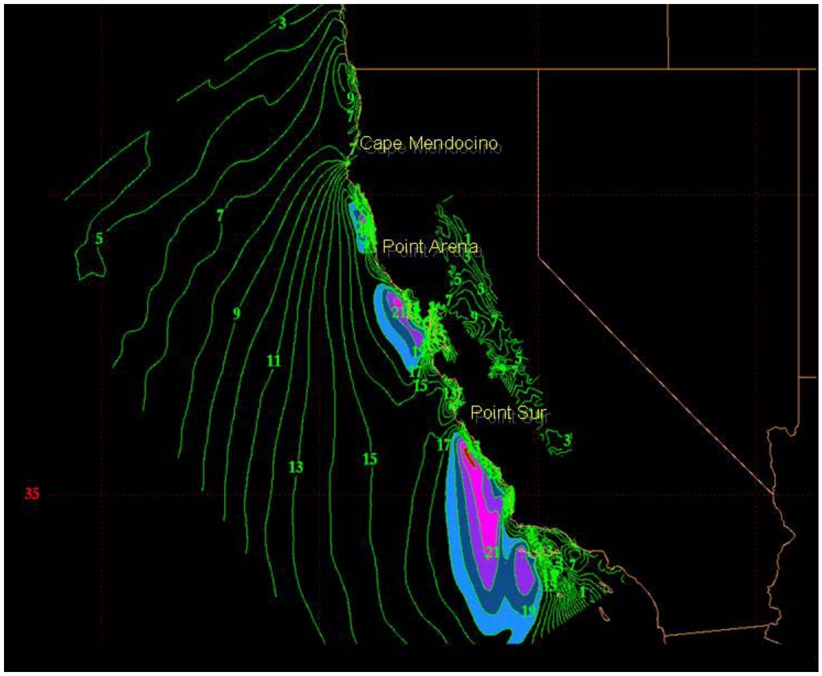


Fig. 5. Wind maximum downwind of Point Arena (COAMPS -- Coupled Ocean/Atmospheric Mesoscale Prediction Systems).

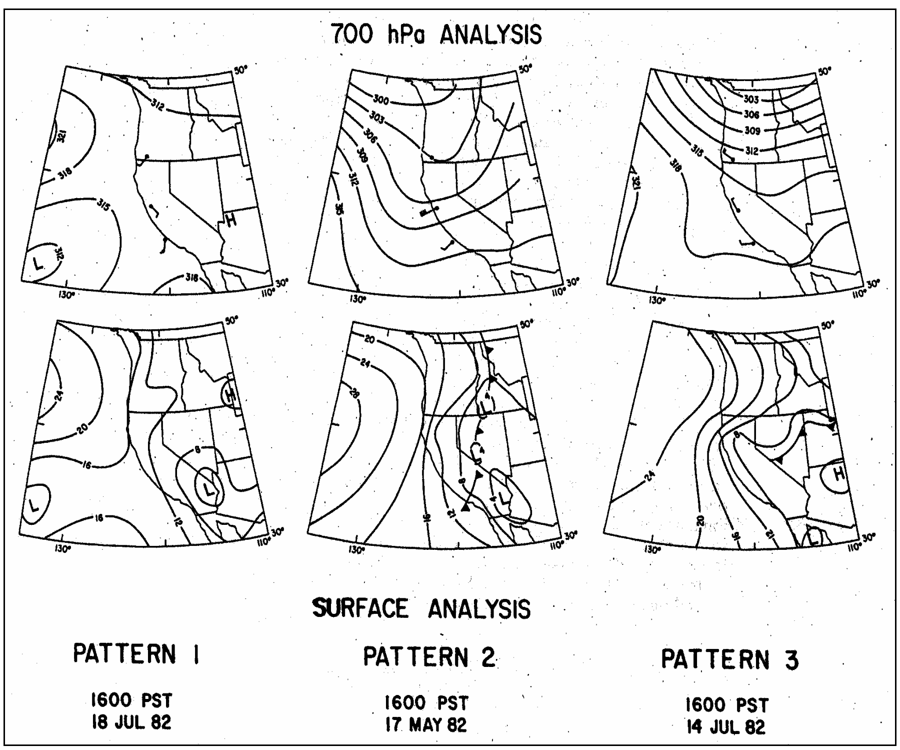


Fig. 6. Synoptic maps at 700 mb and at the surface, which illustrate the three characteristic patterns of the lower atmposphere. (From Winant 1988).

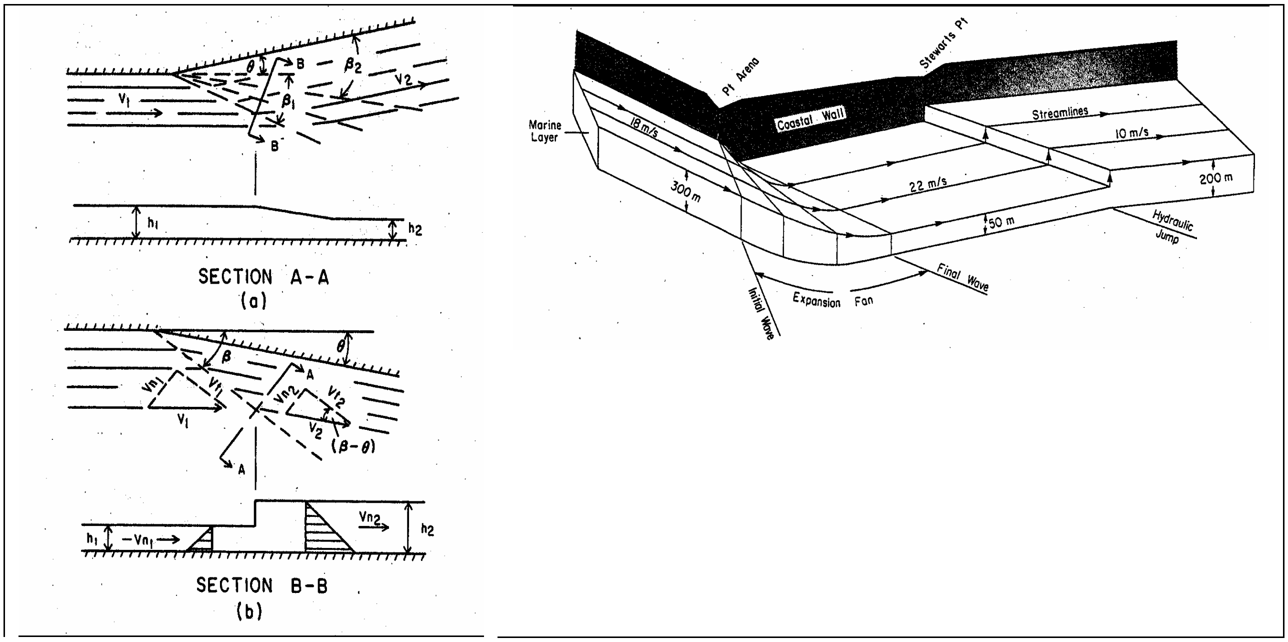
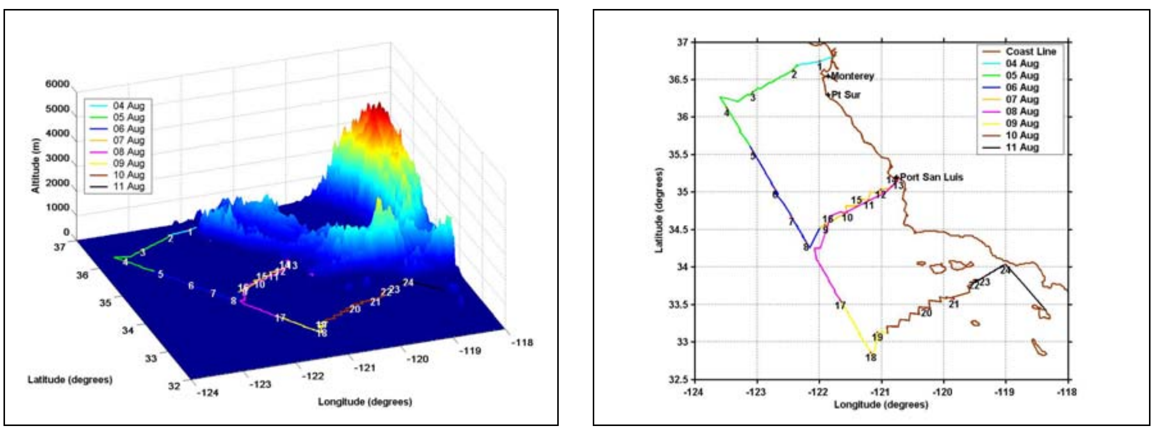
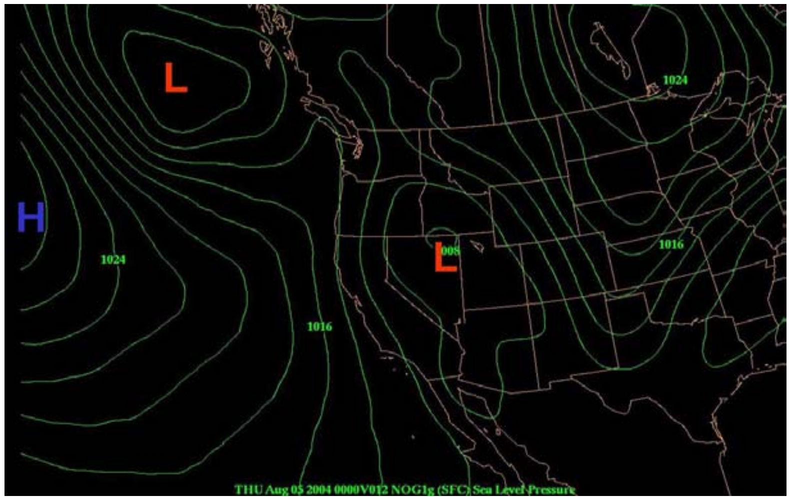


Fig. 7. Definition sketch for (a) supercritical expansion and (b) an oblique jump. (From Winant et al. 1988).



(a) (b)

Fig. 8. Cruise track with the all 24 rawinsonde observations.



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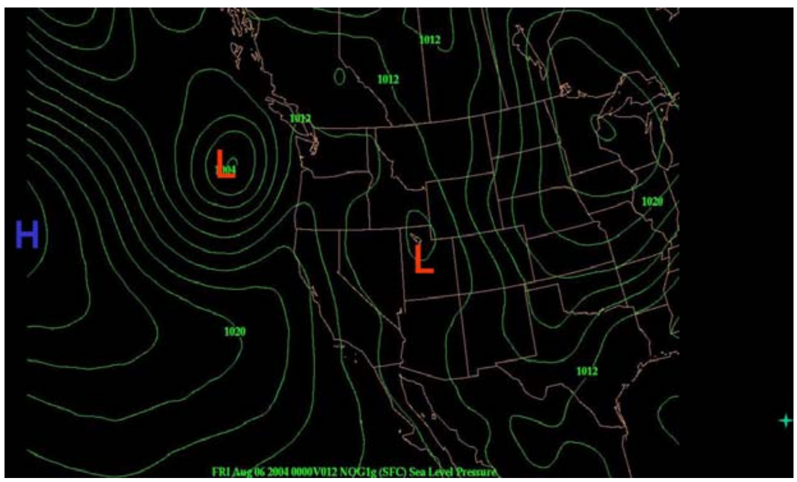

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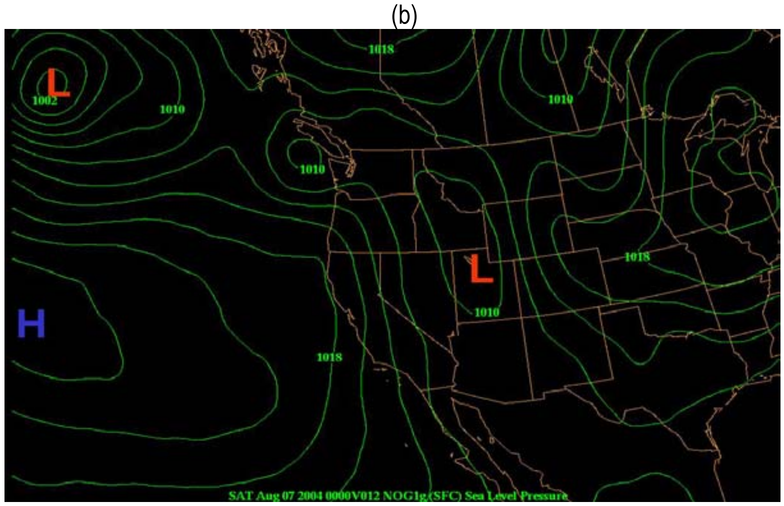
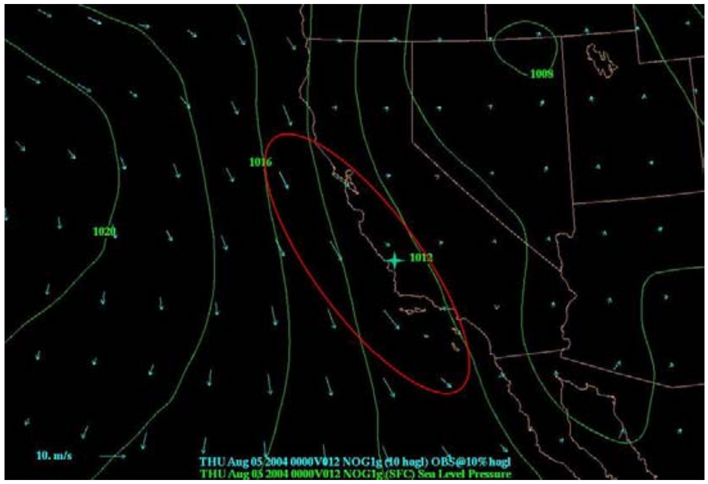


(c)

Fig. 9. NOGAPS synoptic 12-h analysis (surface pressure). (a) 050000UTC, (b) 060000UTC and (c) 070000UTC.



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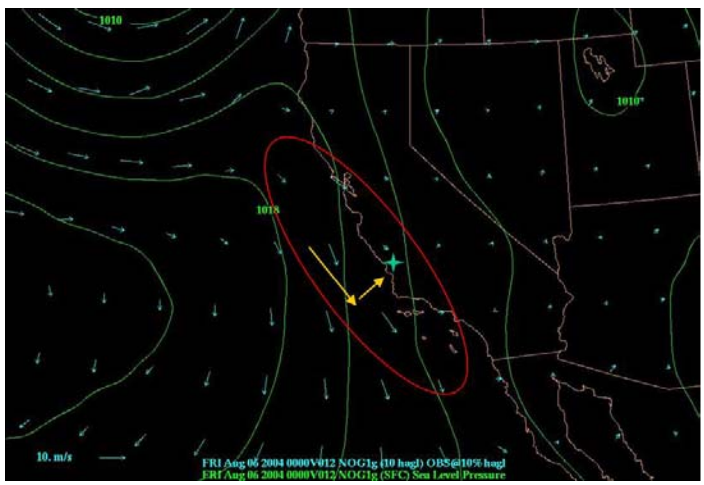

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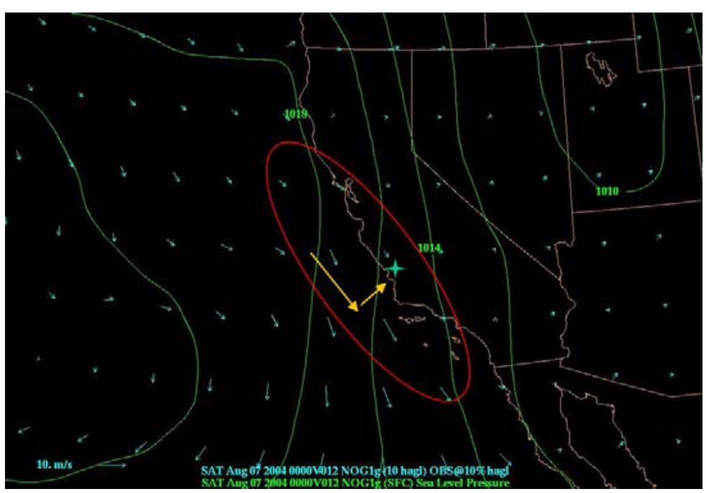
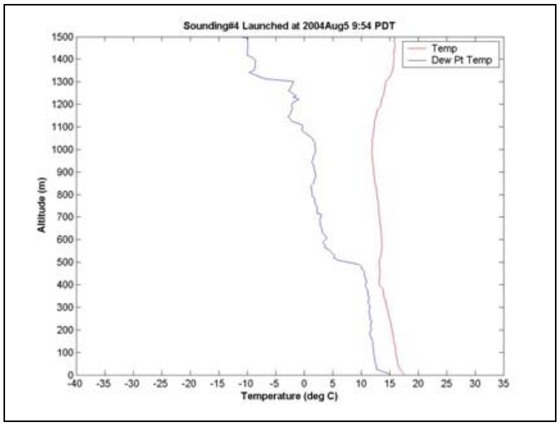


(c)

Fig. 10. NOGAPS synoptic 12-h analysis (surface pressure and 10 wind). (a) 050000UTC, (b) 060000UTC and (c) 070000UTC.



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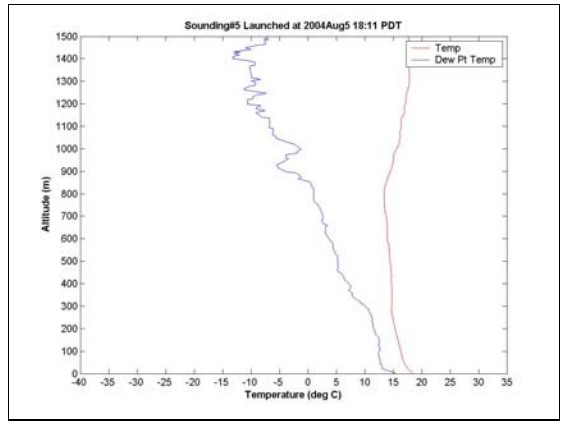

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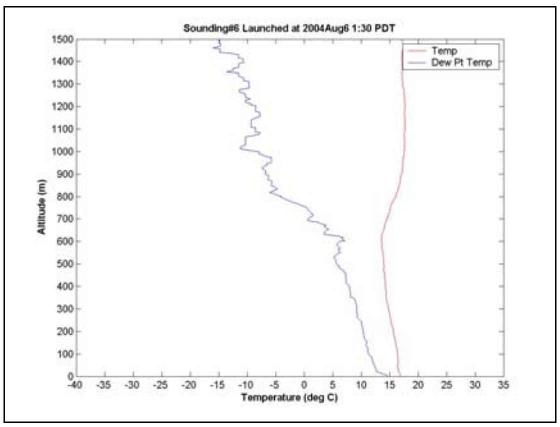

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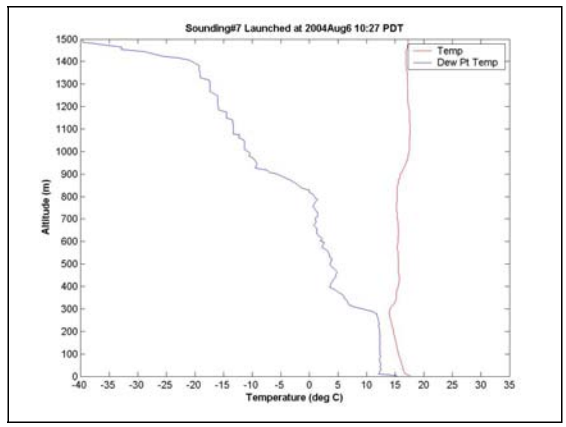

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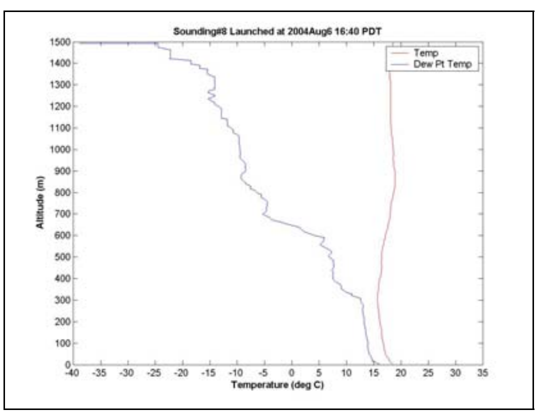

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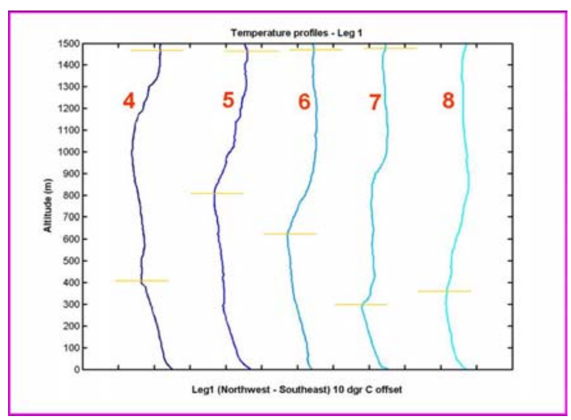

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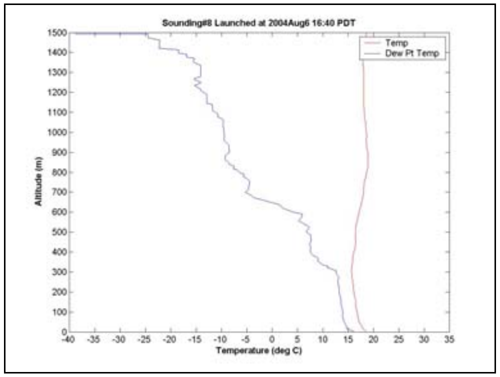
Fig. 11. Temperature and dewpoint profiles (0-1500m) – leg 1.



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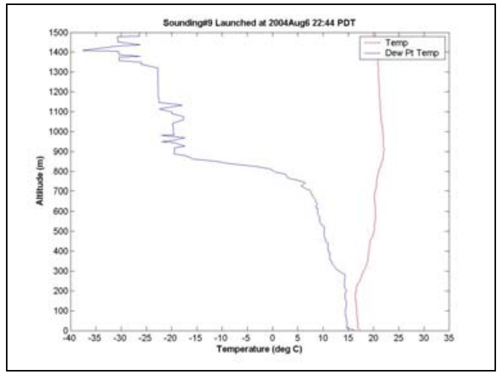

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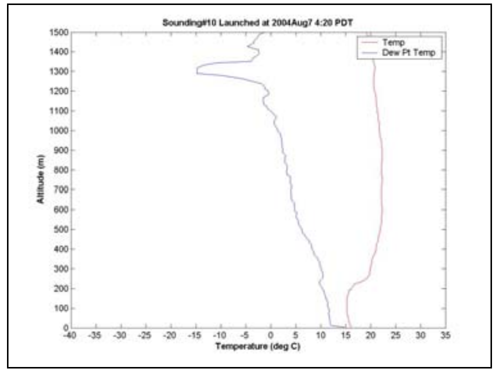

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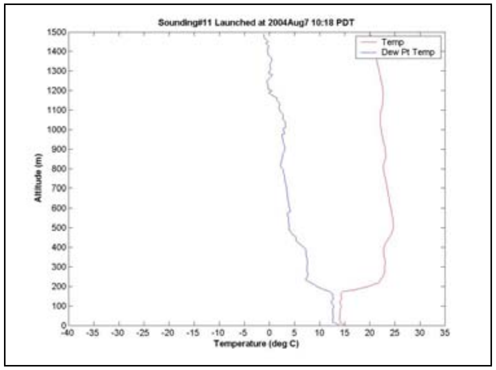

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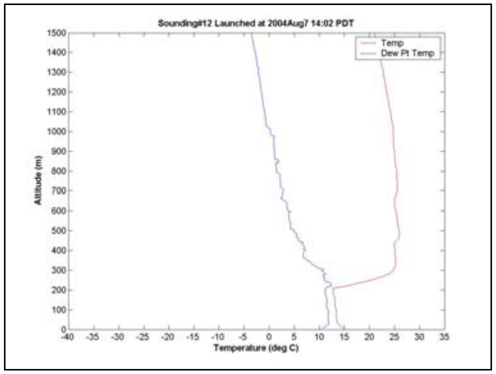

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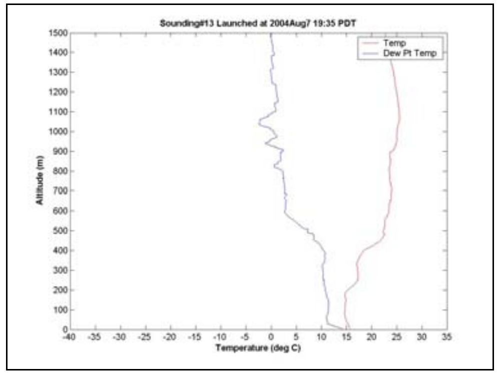
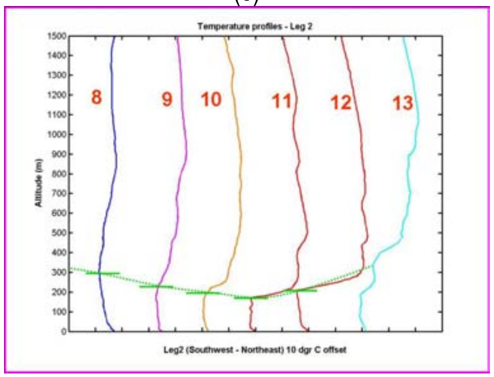
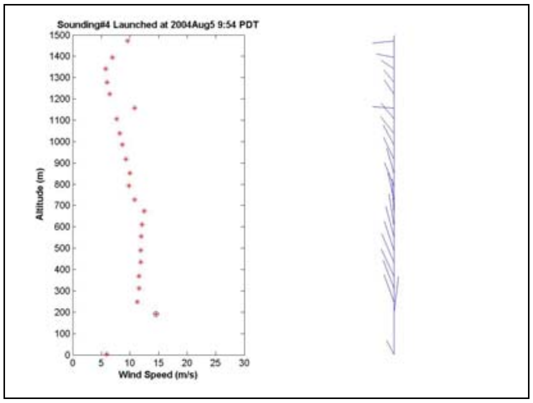


(f)

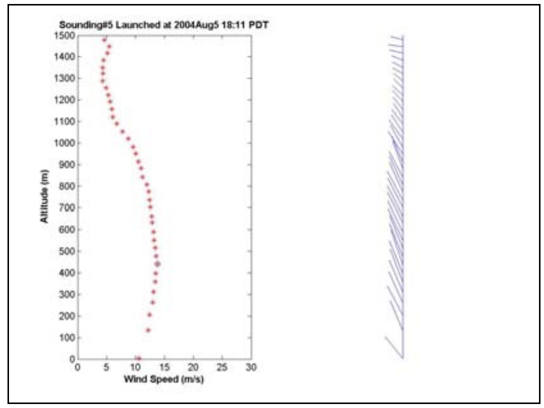


(d)

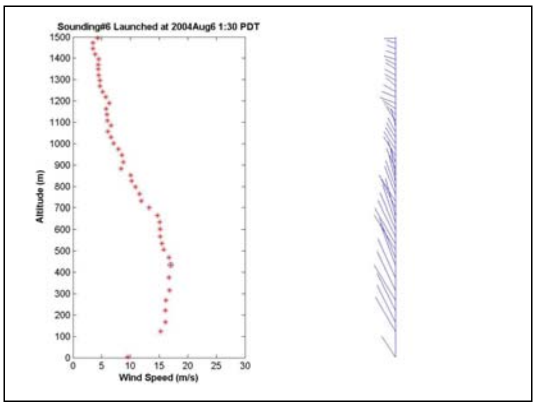
Fig. 12. Temperature and dewpoint profiles (0-1500m) – leg 2.



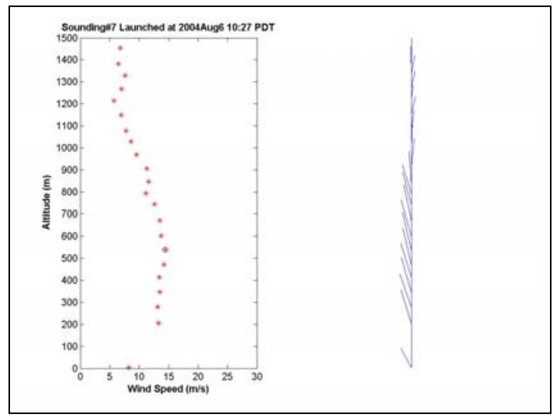
(a)



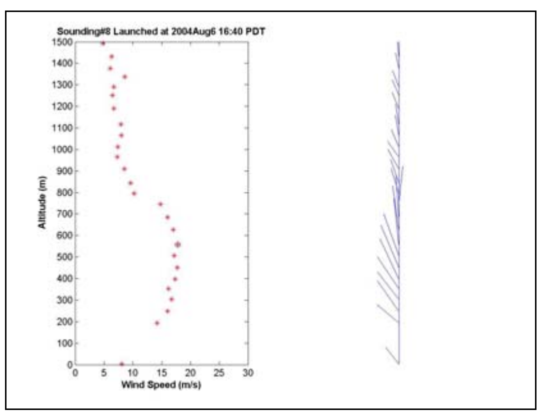
(b)



(c)

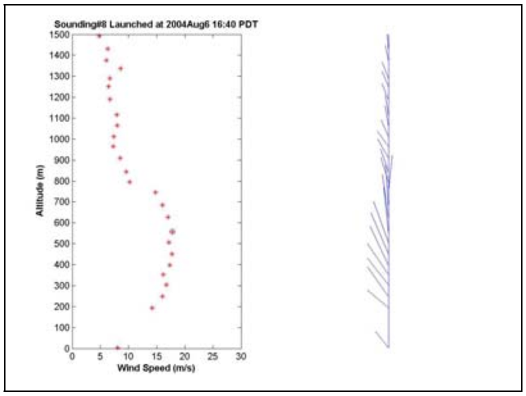


(d)

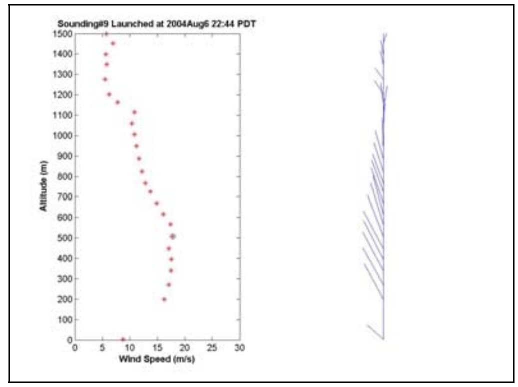


(e)

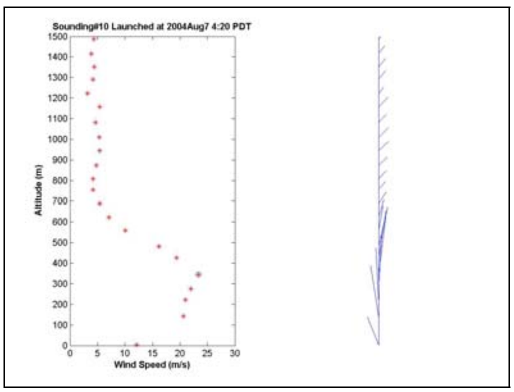
Fig. 13. Wind speed profile and direction (0-1500m) – leg 1.



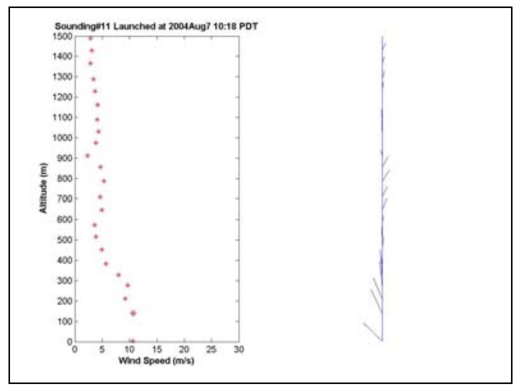
(a)



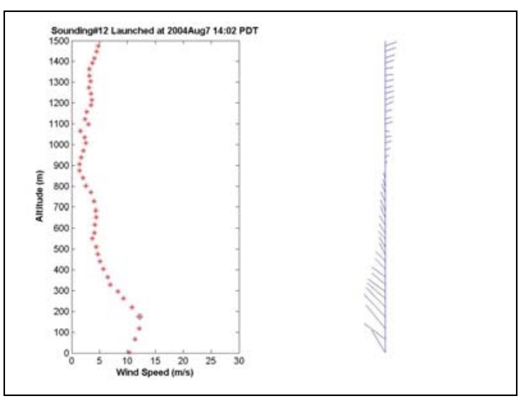
(b)



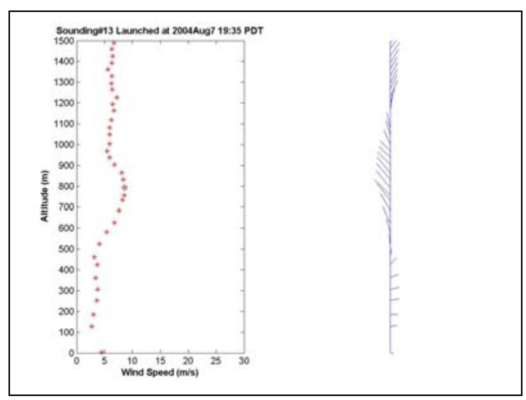
(c)



(d)

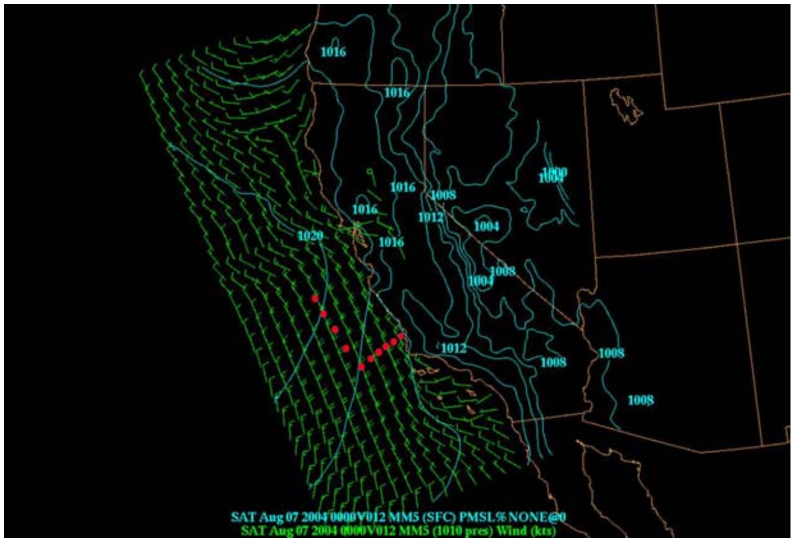


(e)

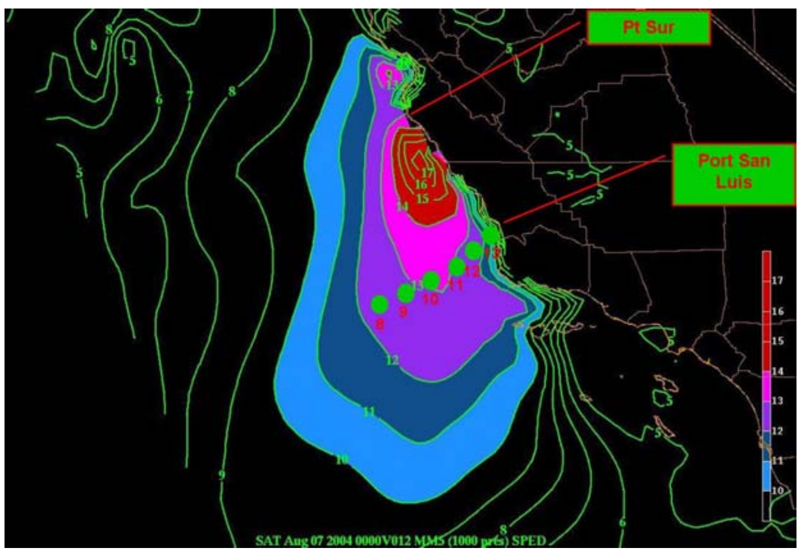


(f)

Fig. 14. Wind speed profile and direction (0-1500m) – leg 2.



(a)



(b)

Fig. 15. MM5 12-h forecast 070000UTC: (a) surface pressure and 24m winds and (b) 1000mb isotachs.

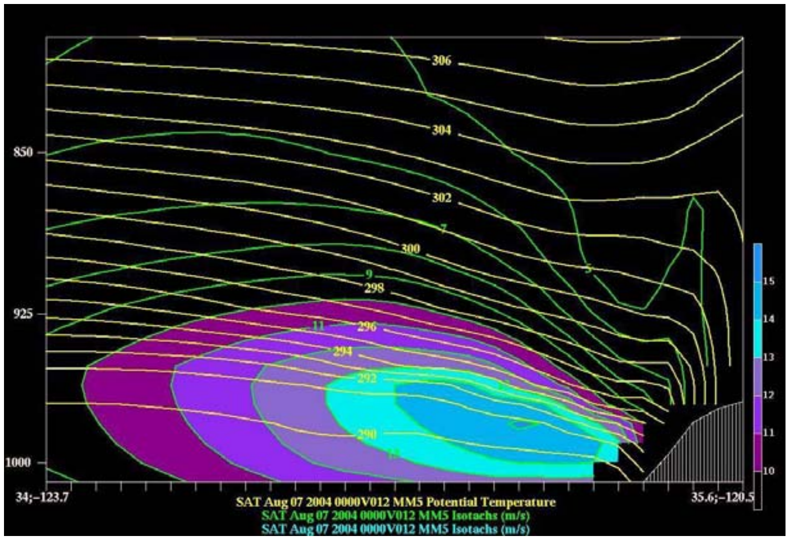


Fig. 16. MM5 12-h forecast 070000UTC: isentropes and isotachs (cross-section along leg 2).

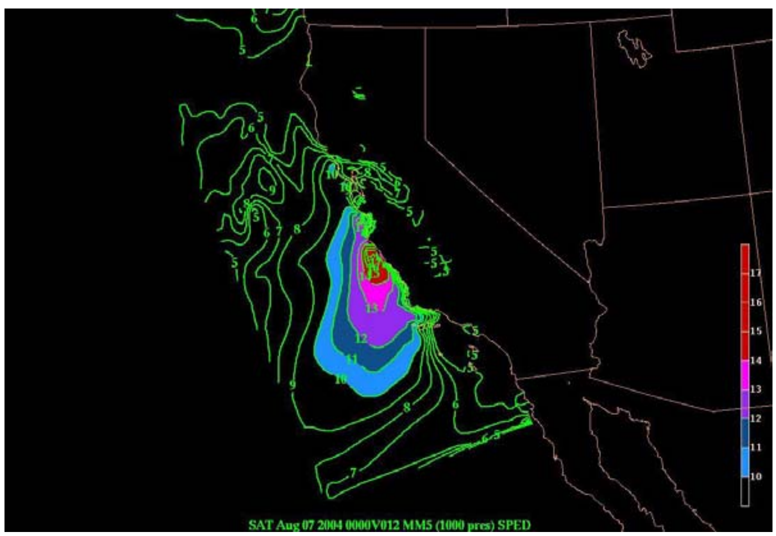


Fig. 17. MM5 12-h forecast 070000UTC: 1000 mb isotachs.

LIST OF REFERENCES

Beardsley, R.C., C.E. Dorman, CA. Friehe, L.K. Rosenfeld, and C.D. Winant, 1987: Local atmospheric forcing during the Coastal Ocean Dynamics Experiment: A description of the marine boundary layer atmospheric conditions over a Northern California upwelling region. *Journal of Geophysical Research*, **92**(C2), 1467-1468.

Burk, S.D., and W.T. Thompson, 1986: The summertime low-level jet and marine boundary layer structure along the California coast. *Monthly Weather Review*, **124**, 668-686.

Chao, S.Y., 1985: Coastal jets in the lower atmosphere. *Journal of Physical Oceanography*, **15**, 361-371.

Cross, P, 2003: The California coastal jet: Synoptic controls and induced mesoscale structure. (Ph.D. Dissertation). *Naval Postgraduate School*.

Dorman, C.E., D.P. Rogers, W.A. Nuss, and Thompson, 1999: Adjustment of the summer marine boundary layer around Point Sur, California. *Monthly Weather Review*, **127**, 2143-2159.

Parish, T., 2000: Forcing of the summertime low-level jet along the California coast. *Journal of Applied Meteorology*, **39**, 2421-2433.

Samelson, R.M., 1992: Supercritical marine-layer flow along a smoothly varying coastline. *Journal of Atmospheric Scien*ce, **49**, 1571-1584.

Winant, C.D., C.E. Dorman, C.A. Friehe, and R.C. Beardsley, 1988: The marine layer off northern California: An example of supercritical channel flow. *Journal of Atmospheric Science*, **45**, 3588-3605.